Location for direct access to subglacial Lake Ellsworth: An assessment of geophysical data and modeling

Citation for published version:

Digital Object Identifier (DOI):
10.1029/2010GL042884

Link:
Link to publication record in Edinburgh Research Explorer

Document Version:
Publisher's PDF, also known as Version of record

Published In:
Geophysical Research Letters

Publisher Rights Statement:
Published in Geophysical Research Letters by the American Geophysical Union (2010)

General rights
Copyright for the publications made accessible via the Edinburgh Research Explorer is retained by the author(s) and / or other copyright owners and it is a condition of accessing these publications that users recognise and abide by the legal requirements associated with these rights.

Take down policy
The University of Edinburgh has made every reasonable effort to ensure that Edinburgh Research Explorer content complies with UK legislation. If you believe that the public display of this file breaches copyright please contact openaccess@ed.ac.uk providing details, and we will remove access to the work immediately and investigate your claim.
Location for direct access to subglacial Lake Ellsworth: An assessment of geophysical data and modeling


Received 10 February 2010; revised 19 April 2010; accepted 28 April 2010; published 2 June 2010.

1. Introduction

[1] Subglacial Lake Ellsworth has been proposed as a candidate for direct measurement and sampling, to identify microbial life and extract sedimentary climate records. We present a detailed characterization of the physiography of this subglacial lake from geophysical surveys, allowing bathymetry and geomorphic setting to be established. Lake Ellsworth is 14.7 km × 3.1 km with an area of 28.9 km². Lake depth increases downlake from 52 m to 156 m, with a water body volume of 1.37 km³. The ice thickness suggests an unusual thermodynamic characteristic, with the critical pressure boundary intersecting the lake. Numerical modeling of water circulation has allowed accretion of basal ice to be estimated. We collate this physiographic and modeling information to confirm that Lake Ellsworth is ideal for direct access and propose an optimal drill site. The likelihood of dissolved gas exchange between the lake and the borehole is also assessed. Citation: Woodward, J., A. M. Smith, N. Ross, M. Thoma, H. F. J. Corr, E. C. King, M. A. King, K. Grosfeld, M. Tranter, and M. J. Siegert (2010), Location for direct access to subglacial Lake Ellsworth: An assessment of geophysical data and modeling, Geophys. Res. Lett., 37, L11501, doi:10.1029/2010GL042884.

2. Geophysical Data Collection

2.1. Radar Surveys

[5] Radio-echo sounding (RES) lines totaling ~740 km were collected in a grid over Lake Ellsworth and the surrounding area (Figure 1) using a ground-based, 1.7 MHz pulsed radar. The ice sheet surface elevation was determined using kinematic GPS with differential correction from the Midlake GPS base station (Figure 1). Ice-bed reflection travel times were converted to ice thickness using a mean radio-velocity in the ice column of 0.168 m ns⁻¹. Estimated RMS errors in ice thickness based on crossover analysis of RES surveys over the lake are ±6.4 m. Surface elevation and ice thickness measurements were then used to construct a Digital Elevation Model (DEM) of the basal topography of the lake catchment (Figure 1).

2.2. Seismic Surveys

[6] Five seismic reflection lines, spaced ~1.4 km apart and aligned perpendicular to the long axis of the lake were completed (Figure 1). With the exception of Profile D, all lines covered the full lake width and part of the surrounding bed. Data were detected using 48 geophones at 10 m spacing. Shot spacing was 240 m, producing single-fold seismic reflection profiles. Data processing included a normal moveout correction and migration.
The ice-base reflection is clear (e.g., Figure 2a) and was picked continuously in all five processed sections. Although the lake-bed reflection could also be picked along most of the lines, short gaps, where the lake-bed reflection could not be picked clearly, were filled by linear interpolation. Similarly, linear interpolation was applied close to the lake edge as the lake-bed reflection could rarely be picked where the water depth is less than \( \sim 10 \) m.

To convert travel-times to layer thickness, mean seismic velocities in the ice column of \( 2846 \) m s\(^{-1}\) (upper 100 m), and \( 3815 \) m s\(^{-1}\) (below 100 m) were used, determined from a short-offset seismic refraction experiment and temperature modeling. Water depths were calculated using a seismic velocity of \( 1437 \) m s\(^{-1}\). Errors in the ice and water column thicknesses (\( \pm 10 \) m and \( \pm 1.5 \) m respectively) arise from uncertainty in the seismic velocities and the travel-time picks. Comparing seismic and RES ice thickness calculations using crossover analysis produces a RMS error of \( \pm 6.3 \) m, showing excellent correspondence between the two data sets.

A DEM of the lake surface was produced from the gridded RES data (Figure 2c). Water column depths calculated from the seismic datasets were then used to produce a DEM of the water column (Figure 2d). To produce this DEM, we interpolated a linear decrease in water depth between the deepest points on the outer seismic lines (A and E) and the corresponding lake end (up-lake or down-lake, respectively). The ice-base and water column DEMs were then combined to give a DEM of bed topography of the lake (Figures 2e).

### 2.3. Ice Flow Measurements

Four GPS receivers were operated on the ice sheet surface to measure ice flow. Three were positioned over the lake (Figure 1) and the fourth was on the adjacent ice sheet \( \sim 11 \) km to the east. Data were recorded at a minimum sampling rate of \( 5 \) s over a period of 60 days. The off-lake station was processed using a kinematic Precise Point Positioning approach implemented in the GIPSY software [King and Aoki, 2003]. The motions of the three on-lake stations were then determined relative to the base station. Henceforth, we will refer to the SE end as up-lake (or upstream) and the NW end as down-lake (or downstream).

### 3. Physiography of Subglacial Lake Ellsworth

#### 3.1. Lake Geometry

Lake Ellsworth sits in the bottom of a narrow, steep-sided valley, is 14.7 km long, has a maximum width of 3.1 km, and a total surface area of 28.9 km\(^2\). The RES data have clarified the shape of the lake’s shoreline (Figure 1). Ice thickness decreases from 3280 m to 2930 m down the lake’s long axis. In the same direction, the lake surface elevation rises from \( -1361 \) m to \( -1030 \) m (Figure 2c). The lake surface and bed profiles shown in Figure 2b reveal a broad, generally U-shaped, lake bed topography. Water depth (Table 1 and Figure 2d) progressively increases down-lake. Maximum water depth on the up-lake line (Line A) is 52 m, increasing to a maximum of 156 m on the down-lake line (Line E).

From the water column DEM (Figure 2d), we calculated a lake volume of \( 1.37 \pm 0.2 \) km\(^3\). The likely uncertainty in this value will be influenced primarily by those parts of the lake where there are no seismic data, upstream of Line A and downstream of Line E. To quantify this uncertainty we considered two extreme situations. 1) The lake bed rises steeply within a short distance of the seismic surveys and the water column then remains shallow (\( \sim 10 \) m) right up to the lake edge. 2) Water depth remains constant away from the seismic lines, with the bed only rising steeply close to the lake edge. These represent extreme, but not wholly unrealistic, possible configurations of lake bathymetry. The shallower case reduces the calcu-
lated lake volume by 0.2 km$^3$, the deeper one increases it by the same amount, giving a conservative error estimate of ±0.2 km$^3$.

### 3.2. Ice Sheet Flow Over the Lake

Ice flow velocities measured with GPS receivers are given in Table 1. The flow speed increases from 4.5 m a$^{-1}$ to 5.5 m a$^{-1}$ from the up-lake GPS station to the down-lake one. Flow direction rotates slightly from 318.9° to 310.8° from up-lake to down-lake.

### 4. Thermodynamics of Subglacial Lake Ellsworth

The bathymetry of Lake Ellsworth suggests an unusual coincidence of thermodynamic characteristics within the lake’s water column. The freezing point of water ($T_f$) and its temperature of maximum density ($T_{md}$) depress at different rates with increasing overburden pressure. These relationships intersect at a critical pressure ($p_c$) of $2.84 \times 10^7$ Pa (Figure 3a) [Wüest and Carmack, 2000]. Where overburden pressure is less than $p_c$, water warmed by geothermal heat flux or latent heat released during freezing is relatively dense and will sink. Conversely, where overburden pressure is greater than $p_c$, warmed water is buoyant and will tend to rise.

Figure 2b shows the location of this critical pressure boundary for each seismic line, converted to depth below the ice surface. For lines A–C, most of the water body sits below the critical pressure boundary. However, for much of the downstream half of the lake, including lines D and E, this boundary lies within the water column. 37% of the lake

### Table 1. Measurements of Profile Widths, Maximum Depths, and Mean Depths For the Five Seismic Cross-Profiles

<table>
<thead>
<tr>
<th>GPS Station</th>
<th>Seismic Line</th>
<th>Width (km)</th>
<th>Maximum Depth (m)</th>
<th>Mean Depth (m)</th>
<th>Ice Flow (m yr$^{-1}$)</th>
<th>Flow Direction (deg)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Offlake</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>1.9 ± 0.1</td>
<td>289.2 ± 2.0</td>
</tr>
<tr>
<td>Uplake</td>
<td>A</td>
<td>1.730</td>
<td>52</td>
<td>26</td>
<td>4.5 ± 0.2</td>
<td>318.9 ± 2.3</td>
</tr>
<tr>
<td></td>
<td>B</td>
<td>2.560</td>
<td>91</td>
<td>43</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Midlake</td>
<td>C</td>
<td>2.900</td>
<td>109</td>
<td>56</td>
<td>4.9 ± 0.1</td>
<td>313.3 ± 0.9</td>
</tr>
<tr>
<td></td>
<td>D</td>
<td>2.795$^b$</td>
<td>143</td>
<td>83</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Lowlake</td>
<td>E</td>
<td>2.665</td>
<td>156</td>
<td>104</td>
<td>5.5 ± 0.1</td>
<td>310.8 ± 0.8</td>
</tr>
</tbody>
</table>

$^a$Ice flow for the off-lake and three on-lake base stations is also shown.

$^b$Seismic profile does not cover width of lake.
volume exists above the critical pressure boundary. This has implications for the circulation model presented below.

5. Ice Water Interaction Within Lake Ellsworth

To investigate the basal mass balance of the ice sheet over Lake Ellsworth we used the 3D numerical fluid-dynamics flow model Rombax [Thoma et al., 2007] to simulate (i) water circulation in the lake; and (ii) the interaction between the water body and the overlying ice. The model has recently been improved by an updated equation of state and a revised equation for the freezing point temperature, according to the Gibbs thermodynamic potential [Thoma et al., 2010]. We assumed that Lake Ellsworth contains fresh water [Vaughan et al., 2007], that the geothermal heat flux is between 46 and 56 mW m$^{-2}$, and the heat flux into the ice is between 20 and 35 mW m$^{-2}$. We applied horizontal and vertical eddy-viscosities of $2 \times 10^4$ m$^2$ s$^{-1}$ and 0.01 cm$^2$ s$^{-1}$ respectively. The model assumes a closed hydrological system (i.e. no water flows into or out of the lake); water enters the lake by melt of the overlying ice and exits by accretion of lake water to the base of the ice sheet. The basal mass balance (Figure 3b) is a direct output of the numerical model. The distribution and thickness of the accreted ice (Figure 3c) however, needs additional information on ice flow at the lake-water interface. The model assumes a constant ice flow velocity of 4.5 m yr$^{-1}$ at the lake surface equal to that measured at the ice surface (i.e. the ice sheet flows across the lake like a floating ice shelf) (Table 1).

The dominant factor driving water circulation is the steep ice-water interface (Figure 2c). Because of this, model results suggest that water circulation occurs within the lake irrespective of the position of the critical pressure boundary. The mean melt rate is $3.8 \pm 0.7$ cm a$^{-1}$ with maximum values of $>16$ cm a$^{-1}$ (Figure 3b). Freezing is predicted for much of the down-lake area, with the highest accretion rates (also $>16$ cm a$^{-1}$) close to the NW shore. Figure 3c shows the steady-state accreted ice thickness, covering $13.3 \pm 0.6$ km$^2$ of the ice-base over the lake (51% of the total surface area). The mean accreted ice thickness is $12.5 \pm 3.5$ cm, reaching a maximum of $\sim 40$ m.

To quantify the model sensitivity to uncertainty in the interpolated water cavity we also ran the model using the two extreme bathymetry grids discussed in Section 3.1. Resulting changes to modeled melting and freezing rates were small (<4%), suggesting that the lack of measurements at the far ends of the lake does not lead to a major difference in the model output.

6. Implications for Direct Access to Lake Ellsworth

The new bathymetry data lead to a number of conclusions regarding the in-situ exploration of Lake Ellsworth. Of primary interest among these is the choice of location where access should be made.

Unlike the conditions found beneath some ice shelves, the basal freezing areas in Lake Ellsworth are unlikely to cause problems for lake access and instrumentation. As modeled freezing and water circulation rates are low, freezing will probably result in the formation of congelation ice (downward ice crystal growth from the ice water interface), rather than frazil ice formation (randomly orientated crystals forming a slush like-layer in the water directly beneath the ice water interface). Initial analysis of seismic reflection strengths shows no systematic difference between the up-lake and down-lake areas, which would be expected if the freezing mechanisms involved frazil ice. The seismic evidence is therefore in accordance with the modeling results. However, until the interpreted melt-freeze characteristics can be confirmed, it would be prudent to position the first entry hole outside our modeled basal freezing area, in case the ice-water interface proves less favorable to access than we interpret. This would suggest drilling should avoid the area downstream of seismic line D.

The hydrologic potential over Lake Ellsworth and its catchment basin [Vaughan et al., 2007] suggests that if there is water through-flow, it will be towards the NW. Accumulation of any terrigenous sediments washed in by subglacial drainage is most likely to occur in the up-lake area where water may enter the lake and deposit its sediment load. These sediments could contaminate or disturb any sedimentary record coming from the ice sheet above the lake, especially if associated with high-energy flood events, so a down-lake location somewhere in the deeper part of the lake is preferred for acquiring a sediment core. Due to lower sedimentation rates, core recovery from a location remote from the source of sediment input is also likely to increase the probability of a longer time series of ice sheet history being recovered.
The ideal location to sample the lake bed sediments would be within a wide, flat bed area, well away from the sides or other steep slopes. The deepest part of the lake basin, where a deep sedimentary record is most likely to be found, is defined by the ~1380 m contour in Figure 2e. The deepest part of the basin sits upstream of the deepest water column and is crossed by seismic lines A to D.

Based on these factors, we propose that the scientific goals for subglacial exploratory research are best achieved at 78°58′4.44″S 90°34′27.56″W (Figures 2 and 3). This location is outside the modeled zone of accretion ice formation, overlies the deepest part of the basin, has deep water (about 143 m), and is away from the up-stream end of the lake where inflow of sediment with basal meltwater may dominate the sedimentary environment.

7. Potential for Gas Build-Up Within the Lake

If we assume that the lake is hydrologically closed there is a potential, over sufficient time, for dissolved gas and clathrates to build up within the lake [McKay et al., 2003]. There is a danger that, during access, this gas may enter the borehole, either as enhanced dissolved gas or as clathrates, and expand to gas as it rises up the borehole, thus causing blowout at the ice surface. Previous experiments that have penetrated subglacial environments in Antarctica e.g. hot water access to the bed of Whillans Ice Stream [Engelhardt et al., 1990] did not report an issue of gasses within the borehole. Subglacial sediments removed from the bed of Whillans Ice Stream did show evidence of degassing [Tulaczyk et al., 2001] suggesting non-hazardous concentrations of dissolved gas.

Furthermore, if SLE was a closed system, dissolved gas concentrations would take ~100,000 years to reach saturation in the lake water at a concentration of ~3.1 cm³/g at 0°C and 1 atm assuming that ice melt in equals ice loss out as accretion ice, with an average melt and freezing rate of 4 cm per year over half the lake, and that accretion ice contains no gas). In this time, the West Antarctic Ice Sheet would have been subject to a full glacial-interglacial cycle, with expansion to a maximum glacial configuration followed by retreat to the present interglacial state. The likelihood of the lake remaining unchanged in size, and in hydrological isolation during this transition is low given the bulk of the ice sheet base is at the pressure melting temperature. We therefore believe the potential of borehole blowout is extremely low.

8. Conclusions

Subglacial Lake Ellsworth reaches a maximum depth of 156 m, has a surface area of 28.9 km² and a volume of 1.37 km³. The combined effects of ice thickness and water column depths suggest an unusual thermodynamic regime, with a change in the sign of thermal expansivity of water occurring within the water column. The steeply sloping ice-water interface determines that water circulation will occur within the lake irrespective of the position of this critical pressure boundary. Assuming the lake to be hydrologically isolated we have modeled the maximum rates of subglacial melting and basal freezeing at the ice water interface; both exceed about 16 cm a⁻¹; and the accreted ice reaches a steady-state thickness of ~40 m.

We conclude that the optimum location to access Lake Ellsworth lies at 78°58′4.44″S 90°34′27.56″W. This reduces the risk from possible basal freezeing mechanisms and optimizes the chances of recovering an undisturbed, continuous sedimentary sequence from the lake floor. This lake access location should allow sampling of the water column to search for microbial life and sampling of the sub-lake sediments to investigate West Antarctic Ice Sheet history.

Acknowledgments. This work was funded by NERC-AFI NE/D008751/1, NE/D009200/1, and NE/D008638/1. We thank BAS for logistics support and NERC-GEF for equipment (loans 838, 870). Dan Fitzgerald and Dave Routledge assisted with the fieldwork.

References


K. Grosfeld, Alfred Wegener Institute for Polar and Marine Research, D-27570 Bremerhaven, Germany.

M. A. King, School of Civil Engineering and Geosciences, Newcastle University, Newcastle-upon-Tyne, NE1 7RU, UK.

M. Thoma, Commission for Glaciology, Bavarian Academy of Sciences, 80539 Munich, Germany.

M. Tranter, School of Geographical Sciences, University of Bristol, Bristol, BS8 1SS, UK.

N. Ross and M. J. Siegert, School of Geosciences, University of Edinburgh, Edinburgh, EH9 3JW, UK.

J. Woodward, School of Applied Sciences, Northumbria University, Newcastle-upon-Tyne, NE1 1ST, UK. (john.woodward@northumbria.ac.uk)